DECIPHERING MARTIAN CLIMATIC HISTORY USING RETURNED

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By necessity, a Mars sample return mission must sample the upper few meters of the Martian surface. This material has been subjected to a wide variety of physical processes. Presently, the most important processes are believed to be wind-driven erosion and deposition, and water ice accumulation at higher latitudes. A sample return mission represents a unique opportunity to better understand and quantify these important geologic processes. By obtaining sample cores at key locations, it may be possible to interpret much of recent Martian climatic history.

It is generally believed that the mid-latitude regions of Mars are presently undergoing net erosion. Surface material raised during great dust storms is transported to the polar regions by the Martian atmosphere and deposited. The ability of the atmosphere to lift and transport dust is a strong function of the Martian surface pressure, which is believed to undergo large variations due to quasi-periodic oscillations in Mars' orbital and axial elements (1). High-resolution Mariner 9 and Viking Orbiter images have revealed extensive layered deposits at both poles, which are believed to contain a sedimentary record of Martian climatic history. These deposits have been estimated to be on the order of 5Km in thickness (2), but the timescale for their formation is not presently clear. Pollack et al. have used dust opacity measurements at the Viking landing sites to estimate dust accumulation rates of approximately 100 meters per million years in the north polar region (3). At this rate of accumulation, the entire column of polar layers could be deposited in 50 million years. Alternatively, if the polar layered deposits accumulated gradually over the entire history of the planet, the accumulation rate would be approximately one meter per million years. Constraining the rates at which the polar layered deposits have been accumulating would be an important step toward understanding martian sedimentary processes and climatic history.

One technique that could be used to determine rates of surface erosion and deposition at low latitudes and in polar terrains involves measuring abundances of short-lived cosmogenic nuclides in near-surface samples. Short-lived isotopes such as ^{10}Be ($\tau_{1/2}$ =1.6My) and ^{26}Al ($\tau_{1/2}$=0.705My) are produced by *in situ* interactions between host rocks and incident cosmic rays. For an isotopically equilibrated rock at constant depth below the surface, the rate of production of short-lived isotopes by cosmic rays is balanced by the rate of loss by radioactive decay. If the depth of rock below the surface changes due to erosion or deposition, then the abundance of short-lived cosmogenic nuclides will not be in equilibrium. The magnitude of the departure from equilibrium depends on the fractional rate of change of nuclide production experienced by the rock (due to a change in depth) relative to the half life of the nuclide. This technique for determining erosion and deposition histories has been successfully applied to lunar samples (4,5) as well as terrestrial rocks (6).

The potential utility of this technique for Martian samples can be investigated by considering the abundances of ^{26}Al and ^{10}Be within dust grains in polar layered deposits. These isotopes are produced primarily by galactic cosmic rays with energies in excess of 100MeV and by secondary neutrons with energies less than 100MeV. ^{26}Al and ^{10}Be in polar layered deposits can be produced by a variety of nuclear reactions arising from cosmic ray-induced spallation of oxygen and silicon. Reedy and Arnold have calculated

production rates for short-lived isotopes as a function of depth below the surface for lunar materials (7). The production rates for these isotopes within dust grains in the polar layered deposits are expected to be an order of magnitude less due to attenuation of cosmic rays by the Martian atmosphere and seasonal CO_2 deposits. Production rates are further reduced by water ice absorption within the layered deposits. Model calculations showing predicted ^{10}Be and ^{26}Al abundances as a function of depth will be presented for a variety of assumptions concerning the composition and depositional history of the layered deposits. Similar calculations could be used to infer depositional rates for actual samples.

A sampling strategy would involve obtaining cores from both high and low latitude locations. Cores at low latitudes are necessary to determine the range of isotopic compositions of polar layered deposit source materials. Cores within the layered deposits should contain a minimum of water ice. Coring depths should be on the order of 100 cm, which is the approximate skin depth for the decrease in production of most cosmogenic nuclides (7). Sample sizes should be large enough to obtain sufficient analytical precision to distinguish between model erosional and depositional history scenarios. We calculate that a core diameter of 1 centimeter and a typical dust layer thickness of 2 microns could yield enough sample to determine ^{26}Al and ^{10}Be abundances to an accuracy of 2% using conventional accelerator mass spectrometric techniques (8). This precision would be sufficient to infer erosion or deposition rates ranging from 1 cm per million years to 10 Km per million years. Therefore, these measurements could place important constraints on our understanding of Martian sedimentary processes and climatic history.

REFERENCES

1. Ward W.R. (1974) *J. Geophys. Res.* **79**, 3375-3395.
2. Dzurisin D. and Blasius K.R. (1975) *J. Geophys. Res.* **80**, 3286-3306.
3. Pollack J.B., Colburn D.S., Flaser F.M., Kahn R., Carlston C.E. and Pidek D. (1979) *J. Geophys. Res.* **84**, 2929-2945.
4. S.H.R.E.L.L.D.A.L.F.F. (1970) *Proc. Apollo 11 Lunar Sci. Conf., Geochim. Cosmochim. Acta* **1**, 1503-1532.
5. Finkel R.C., Aronld J.R., Imamura M., Reedy R.C., Fruchter J.S., Loosli H.H., Evans J.C., Delany A.C and Shedlovsky J.P. (1971) *Proc. 2nd Lunar Sci. Conf.* **2**, 1773-1789.
6. Nishiizumi K., Lal D., Klein J., Middleton R. and Arnold J.R. (1986) *Nature* **319**, 134-136.
7. Reedy R.C. and Arnold J.R. (1972) *J. Geophys. Res.* **77**, 537-555.
8. Elmore D. and Phillips F.M. (1987) *Science* **236**, 543-550.